

northwestern Laurentide ice sheet would also have contributed to some of the East Greenland Current discharge. A preliminary evaluation of Mediterranean $\delta^{18}\text{O}$ records points to only a minor contribution of glacial meltwater from rivers draining into the Black Sea, in agreement with the conclusions of other workers⁴⁰. The volume estimates contained in Table 2 should be considered minimum estimates, however, because potential contributions of meltwater from the Arctic Ocean through the Bering Strait (after its submergence) are not included in this analysis.

Simple calculations using the volume estimates in Table 2 place limits on the minimum $\delta^{18}\text{O}$ of the ice sheets (Table 3) for a probable range of values suggested as representing the glacial-interglacial change in $\delta^{18}\text{O}_{\text{sw}}$ (ref. 41). If significant meltwater contributions can be traced to the Bering Strait and to Antarctic deglaciation, the values in Table 3 would be increased proportionately (made less negative). The mean $\delta^{18}\text{O}$ of the ice sheets cannot be taken to represent the average $\delta^{18}\text{O}$ of the meltwater dilutant during a specific time interval, however, because of possible large variations in $\delta^{18}\text{O}$ within the ice sheets³⁸ and unknown variations in the $\delta^{18}\text{O}$ of precipitation during deglaciation. While we are therefore at present unable to accurately estimate salinity from estimated $\delta^{18}\text{O}_{\text{sw}}$, we feel the approach adopted here makes it possible to better understand the dynamics of meltwater discharge during the last, and possibly previous, deglaciations of the Pleistocene.

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Sulphur-rich volcanic eruptions and stratospheric aerosols

Michael R. Rampino*[†] & Stephen Self[‡]

* NASA Goddard Institute for Space Studies, New York, New York 10025, USA

† Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964, USA

‡ Department of Geology, University of Texas at Arlington, Arlington, Texas 76019, USA

During the past decade it has become clear that the long-lived stratospheric clouds produced by volcanic eruptions are composed largely of sulphuric acid aerosols^{1,2}. The amount of sulphur-rich volatiles (for example, SO_2 , H_2S) injected into the stratosphere by an explosive eruption is, therefore, a critical determinant of its atmospheric impact^{3,4}. The small-volume eruptions of Mt Agung in 1963 and El Chichón in 1982 both generated substantial stratospheric aerosol clouds, despite the fact that they erupted $\leq 0.5 \text{ km}^3$ of magma. Comparison of data from direct measurements of stratospheric optical depth, Greenland ice-core acidity, and volcanological studies shows that such relatively small, but sulphur-rich, eruptions can have atmospheric effects equal to or even greater than much larger sulphur-poor eruptions. These small eruptions are probably the most frequent cause of increased stratospheric aerosols.

The recent eruption of El Chichón has provided fresh evidence that small-volume explosive events can have significant atmospheric impact^{5,6}. An older, but classic example is the 1963 eruption of Mt Agung (Gunung Agung) on the island of Bali, which is considered to be among the most important volcanic events of the twentieth century, primarily because of its possible effects on global climate^{7,8}. The 1963 Agung eruption (8°S) was a small-volume vulcanian to subplinian explosive event that produced about $0.3\text{--}0.6 \text{ km}^3$ of basaltic andesite magma. It has been given a volcanic explosivity index (VEI)^{9,10} rating of 4 on the scale of 0 to 7. The VEI is a combined measure of the explosive violence and volume of discharged pyroclastics of an eruption. The rather unusual Agung eruption sequence, with an early lava flow followed by climactic explosions, may be explained by the mixing of two distinct magma types¹¹. The duration of the major explosive phase on 17 March 1963 was about 7 h, while the 16 May eruption lasted for 5 h, suggesting peak rates of output of dense magma of $\sim 10^4 \text{ m}^3 \text{ s}^{-1}$. This is calculated to have produced eruption columns at least 18 km above sea level (a.s.l.)¹². The eyewitness reports of column heights on 17 March and 16 May were only 13 km a.s.l.¹³, but ash and sulphate aerosols were collected over Australia at 20 km a.s.l. from April 1963 up to about a year afterwards¹⁴. The resultant change in stratospheric aerosol optical depth¹⁵ peaked in the Southern Hemisphere subtropics and mid-latitudes at 0.2 to 0.3 from August to November 1963; Northern Hemisphere peak optical depths were much less, only ~ 0.02 measured at Hawaii¹⁶ during late 1963.

The measured optical-depth changes after the eruption suggest $\sim 1\text{--}2 \times 10^{13} \text{ g}$ of aerosol in the stratosphere^{17,18}. Based on the maximum estimated bulk volume of Agung ashfall ($\sim 1 \text{ km}^3$), the total amount of very fine ($< 2 \mu\text{m}$) ash that could have been produced by the eruption is roughly estimated¹⁹ as $< 8 \times 10^{12} \text{ g}$. Direct high-altitude sampling soon after the eruption suggests $\sim 5 \times 10^{12} \text{ g}$ of ash in the stratosphere²⁰. Such fine ash has only a brief stratospheric residence time, however, and was largely removed within a few months, leaving an aerosol cloud composed mainly of H_2SO_4 droplets²⁰.

For comparison, the El Chichón (17°N) eruption in 1982 was also small in volume, producing about $0.3\text{--}0.35 \text{ km}^3$ of trachyan-desite magma^{21,22}. The eruption had three main subplinian explosive phases of VEI = 4 on 28 March, 3 and 4 April. The eruption columns penetrated the stratosphere, where the main

Table 1 Comparison of eruption characteristics

Eruption	VEI ⁹	Volume of magma erupted (km ³)	Eruption column height (km) (c = calculated o = observed)	Stratospheric H ₂ SO ₄ aerosols (g) (from optical depth) ¹⁷	Stratospheric H ₂ SO ₄ aerosols (g) (from ice-core acidity) ²⁹	H ₂ SO ₄ yield (g) (from inclusion analysis) ²⁸	Northern Hemisphere ΔT (°C)	Refs (cols 3–8)
Tambora (8° S) 1815	7	≥50	>40 (c)	2×10^{14}	1.5×10^{14}	5×10^{13}	-0.4 to -0.7	35, 35, 17, 29, 28, 17
Krakatau (6° S) 1883	6	≥10	>40 (c)	5×10^{13}	5.5×10^{13}	3×10^{12}	-0.3	3, 3, 1, 29, 28, 36
Santa Maria (15° N) 1902†	6	~9	>30 (c, o)	$\leq 2 \times 10^{13}$	$\leq 2 \times 10^{13}$	—	-0.4	37, 37, †, †, —, 36
Katmai (58° N) 1912	6	15	>27 (c)	$\leq 2 \times 10^{13}$	$\leq 3 \times 10^{13}$	—	-0.2	38, 38, †, †, —, 36
Mt St Helens (46° N) 1980	5	0.35	22 (o)	$\sim 3 \times 10^{11}$	—	8×10^{10}	0 to -0.1	39, 40, 40, —, 28, 41
Agung (8° S) 1963	4	0.3 to 0.6	>18 (c) 13 (o)	$1-2 \times 10^{13}$	‡	3×10^{12}	-0.3	11, 11, 23, —, 28, 36
El Chichón (17° N) 1982	4*	0.3 to 0.35	26 (o)	$1-2 \times 10^{13}$	—	7×10^{10}	-0.4 to -0.6	22, 6, 23, —, 28, 42
Laki (64° N) 1783	4	0.3 (tephra) 12.3 (lava)	?	—	$\leq 1 \times 10^{14}$	9×10^{13}	~-1.0	4, —, —, †, 28, 43

* S. S., unpublished data.

† See text.

‡ Hammer *et al.*²⁹ give 2×10^{13} g, but this acidity peak cannot be a result of the Agung eruption, see text.|| Jones⁴² states that much of this decrease occurred between January and March 1982, before the eruption took place.

stratum of El Chichón aerosol spread out at about 26 km altitude⁶. Estimates based on lunar eclipse data²³ give an average global peak optical depth of 0.12 in December 1982; the aerosol was preferentially concentrated in the Northern Hemisphere, where the optical depth was 0.15. Conversion from optical depth to sulphuric aerosol mass loading¹⁷ yields about 2×10^{13} g of H₂SO₄ aerosol in the global stratosphere 6 months after the eruption. This agrees with estimates based on airborne lidar measurements²⁴ (1.2×10^{13} g) and on balloon-borne particle counters²⁵ ($1-2 \times 10^{13}$ g).

The large amount of sulphur volatiles released during these eruptions and injected into the stratosphere seem to be the most critical factor in their long-term atmospheric effects^{3,4}. Several studies have suggested that iron-rich basaltic and andesitic melts have a generally high capacity to carry dissolved sulphur^{4,26}. Melts of iron-poor dacite to rhyolite composition that commonly produce large volume, explosive plinian and ignimbrite eruptions (for example Krakatau 1883, Santa Maria, 1902) are usually poorer in sulphur volatiles^{3,27,28}.

Table 1 compares the vital statistics of significant explosive volcanic eruptions of the past 200 yr. The explosive magnitude of an eruption is succinctly summarized by its VEI value (column 2). The largest eruptions produced voluminous airflow and pyroclastic flow deposits (ignimbrite) (column 3) and had very high eruption columns (column 4). The global stratospheric aerosol mass loadings in column 5 were calculated using the method of Stothers¹⁷, where mass loadings (M_D) may be obtained from the global-average peak optical depth measurement (τ_D) using the formula $M_D = 1.5 \times 10^{14} \tau_D$ (grams). The estimates of global stratospheric aerosols based on Greenland ice-core acidities (column 6) are from Ref. 29. The values for Santa Maria, Katmai and Laki are given as upper limits because the aerosol clouds from these eruptions were probably restricted to the Northern Hemisphere or some portion of it. The sulphuric acid yield of the eruptions listed in column 7, taken from Ref. 28 are based on microprobe analyses of sulphur in volcanic ash samples and on estimates of the pre-eruption sulphur content of the magma from analyses of sulphur volatiles in glass inclusions trapped within crystals that formed just before the eruption²⁸. The values in column 7 were derived by subtracting the amount of sulphur remaining within the ash from the amount of sulphur volatiles found in the inclusions and then multiplying by the total volume of erupted magma. This 'sulphur yield' was then converted into the equivalent sulphuric acid aerosol mass²⁸.

The maximum observed post-eruption Northern Hemisphere temperature decreases that might be associated with the erup-

tions are listed in column 8. Specifically, ΔT is the difference between the temperature in the year before the eruption and the lowest temperature in the 1–3 yr following the eruption. In the case of St Helens, Northern Hemisphere temperatures rose in the 2 yr following the eruption, and the quoted decrease is the theoretical value of Robock⁴¹. However, not all of the temperature decreases can be attributed to the volcanic aerosols; in several instances the temperature decreases began before the eruption in question and in all cases, the signal is of the same order as interannual variability⁷. Furthermore, in certain cases, (for example, 1902) several eruptions capable of producing stratospheric aerosols may have occurred in close succession [for example, Pelée 1902 (VEI = 4) and Soufrière 1902 (VEI = 4)].

Several conclusions can be reached based on the data in Table 1. First, small magnitude volcanic eruptions can produce atmospheric after-effects like those of much larger eruptions. Second, there is good agreement between the aerosol estimates based on direct observations of changes in stratospheric optical depth^{3,17,18,23}, and those based on acid fallout over the Greenland ice sheet²⁹. In the case of Agung, however, the agreement must be fortuitous; stratospheric optical depth measurements clearly show that the bulk of the aerosol stayed in the Southern Hemisphere¹⁵. The small Northern Hemisphere optical depth of 0.02 indicates only $\sim 3 \times 10^{12}$ g of aerosol in the Northern Hemisphere stratosphere. The Greenland acidity peak in 1963–64 is too large, possibly because of input of aerosols from high northern latitude eruptions (such as, Surtsey in Iceland). Third, in all cases, the calculations of total sulphur release based on volcanological studies²⁸ are evidently underestimates, although in a relative sense they are in general agreement with the other methods. For El Chichón, the volcanological estimate of sulphur release is more than 2 orders of magnitude too low compared with the direct measurements of sulphuric acid aerosols in the stratosphere. Clearly these analyses are missing some source of additional sulphur release.

Where could the excess sulphur released in these eruptions have come from? In the case of El Chichón, anhydrite (CaSO₄; considered rare in volcanic rocks) occurs as a common phenocryst phase in the deposits, and decomposition of anhydrite has been suggested as a source of additional sulphur during the eruption^{22,28}. The Agung deposits, however, show no evidence of anhydrite. Another possible source of sulphur volatiles is degassing of non-erupted magma³⁰. Rose *et al.*^{30,31} calculated that for the 1974 eruption of Fuego (Guatemala) (also a small andesitic eruption) the amount of sulphur released required degassing of a body of intrusive magma about 5 times the mass

of the erupted material. If a similar situation occurred during the Agung eruption, then the total sulphur release would have been about 2.6×10^{12} g (or enough to produce about 8×10^{12} g of H_2SO_4 aerosols).

This study suggests that frequent, very-intense vulcanian to sub-plinian eruptions from the world's many active basaltic-andesitic composite volcanoes, such as Agung 1963, are probably the most common cause of significant perturbations of the stratospheric aerosol layer. These small eruptions are often rich in sulphur volatiles³¹ and can inject them efficiently into the stratospheric aerosol layer. The source of the sulphur enrichment, however, is not obvious. To solve this problem, a better understanding is required of the sulphur systematics in magmas and the processes taking place in shallow magma bodies beneath the volcanoes. For example, physical mixing of magmas of different composition may have been an important driving force in the Agung eruption and in many other basaltic-andesite volcanoes³²; such magma-mixing events may also control the concentrations and mode of release of sulphur gases and other volatiles during these eruptions.

A possible connection has been previously established between volcanic perturbations of the stratospheric aerosol layer and short-term (1–5 yr) climate cooling^{3,4,7,33,34}. We note from our new results in Table 1 that for the eruptions studied, ΔT is roughly proportional to $M_D^{1/3}$, as has been found previously by Devine *et al.*²⁸. Studies of polar ice cores show a further correlation between longer (decadal) periods of elevated ice-core acidity and cool climate²⁹. This suggests that episodes of increased sulphur-rich explosive volcanism may be an important modulator of long-term climate change. The connection between sulphur-rich volcanism and climate cooling may extend to large-volume basaltic fissure-type eruptions such as Laki (1783)^{4,28}. Large quantities of sulphuric acid aerosols could possibly be lofted into the lower stratosphere by such eruptions²⁸; ultimately, flood-basalt extrusions may have the greatest effects on the global atmosphere of any volcanic events.

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On the formation of calderas during ignimbrite eruptions

T. H. Druitt* & R. S. J. Sparks

Department of Earth Sciences, University of Cambridge, Cambridge CB2 3EQ, UK

Many large calderas result from the eruption of substantial volumes (tens or hundreds of km^3) of silicic pyroclastics. Such events often begin with an airfall phase and progress to the generation of voluminous ignimbrites^{1–3}. We propose here that many such eruptions involve two well-defined stages, based on a simple analysis of magma chamber pressure variations during an eruption. The first stage begins when an overpressured magma chamber fractures the country rock and forms a conduit to the surface. The chamber pressure decreases rapidly to values less than lithostatic pressure. We show that only small to moderate volumes of magma, representing a small fraction of the total chamber, can be erupted during this stage. In the second stage, caldera collapse results from a further decrease in magma pressure, which causes the chamber roof to fracture catastrophically and deform. Subsidence of the roof attempts to re-establish lithostatic pressures within the chamber and can drive substantial volumes of magma to the surface. Geological relationships in pyroclastic deposits associated with large caldera eruptions provide independent evidence for this model.

Present ideas on caldera formation are strongly influenced by studies in which it was recognized^{1,4} that collapse was fundamentally the consequence of removal of magma from the chamber. Smith and Bailey⁴ presented a model of caldera evolution, based largely on their studies of the Valles Caldera, New Mexico. They envisaged that an eruption began when a large body magma had accumulated within the upper crust, causing regional uplift and stretching. They also recognized that collapse often occurred while large volumes of ignimbrite were erupted from the chamber. The area of collapse is often proportional to the volume of erupted magma², suggesting a simple relationship between caldera collapse and removal of magma from the chamber.

The broad features of these classic ideas are widely accepted and are not in dispute here. We now present a simple analysis of magma chamber pressure during eruption, which suggests that caldera-forming events often have two distinct phases.

Figure 1 illustrates the two stages which are envisaged. Before eruption, magmatic overpressure becomes large enough to support the chamber roof and to propagate tensile or shear fractures to the surface^{4,5}, triggering an eruption. Smith and Bailey proposed that concentric, inward-dipping ring fractures were formed during the pre-eruption doming, and that these fractures provided pathways for initial magma ascent to the surface. Although there is evidence for early ring fracture development in some calderas^{6,7}, studies of the distribution of initial airfall deposits around volcanic centres^{1,8–13} show that eruptions can also begin in localized areas, perhaps along regional faults, which are not obviously connected to a ring fracture system. In

* Present address: US Geological Survey, MS-910, Menlo Park, California 94025, USA.